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NASA TECHNICAL TRANSLATION NASA TT F-14,501

METHOD FOR DETERMINATION OF THE HEAT FLUX IN SOIL AND ITS APPLICATION TO FIELD MEASUREMENTS

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Translation of "A talajelvezetési tényező meghatározásanak módszere és alkalmazása terepen mért adatokra," Idojaras, Vol. 64, No. 6, 1960, pp. 365-371.

NATIONAL AERONAUTICS AND SPACE ADMINISTRATION
WASHINGTON, D. C. 20546 AUGUST 1972

METHOD FOR DETERMINATION OF THE HEAT FLUX IN SOIL AND ITS APPLICATION TO FIELD MEASUREMENTS

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ABSTRACT: The surface energy balance is discussed briefly and is followed by an analysis of the thermal and physical characteristics of the soil. Tseytin's method for determining the heat flux in the soil is discussed in detail and the method is applied to data from measurements made at Lake Balaton in 1959. The results are compared with data for other heat balance factors.

Knowledge of the heat balance for different active surfaces requires /365* determination of the value (Q_t) of the soil's heat flux. This factor indicates the quantities of heat entering and/or leaving the soil. The overwhelming percentage of the energy arriving from the sun exerts its effects by the intervention of the soil. Thus, the soil influences the condition of the above- /366 ground air layers according to its heat balance characteristics. The surface soil transmits thermal energy to the atmosphere, and to the lower strata of the soil by conduction. The atmosphere is a much poorer heat conductor than the soil (the conductivity of dry air is 0.00005 cal/cm-degree-sec, that of sand is 0.003, that of loose rock 0.004, and that of the wet marshy soil 0.002 cal/cm-degree-sec). Thus, surface soil transmits much more of the energy in its heat reserve to the lower layers of the soil than to the adjacent atmosphere. On the other hand, the heat lost by the surface by radiation is all into the air. Heat loss also accompanies evaporation of water from the surface. Conversely, condensation increases the surface heat reserve because it liberates only as much heat as the evaporation of water needs.

The 24-hour surface heat balance can be broken down into a daytime, so-called irradiation phase, and a nighttime, so-called radiation phase. The surface warms markedly during the day, developing a downward temperature gradient, and causing heat to penetrate the deeper layers. At night the situation is the reverse. The surface cools as a result of radiation and the direction of the gradient is reversed, causing heat to rise and the soil to cool. Daytime irradiation generally is stronger from spring to late autumn, so the heat intake

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* Numbers in the margin indicate pagination in the foreign text.

by the soil is greater than the heat loss by radiation. The daily heat exchange ends with a gain, and this gain is used to warm the deeper soil layers. On the other hand, the opposite is true from late autumn to spring. Heat stored in the deeper layers rises to the temperature of the upper layers that now are getting colder because of more intensive radiation.

The heat balance properties of individual soils are greatly influenced by the soil's physical characteristics and condition, including the capacity to absorb radiation, thermal conductivity, specific heat, thermal capacity, and temperature conductivity [1].

Only the capacity of the surface to absorb radiation plays a role in the soil's heat balance because most of the radiation already is absorbed in the upper layers a few millimeters in thickness. The darker the soil, the greater the absorption of radiation.

The thermal conductivity of the soil characterizes the rate of heat penetration into the soil. Soil containing a great deal of air is a poorer conductor of heat than a more compact soil because, as we already have mentioned, air is a very poor heat conductor.

The specific heat is a very important characteristic of soil physics. We know that the specific heat of the mineral components of the soil is almost identical, but that the specific heat of the organic components is twice larger, and that of water is five times larger. It follows therefore, that the specific heat of soil is highly influenced by its moisture content, so that the damper the soil, the less it is warmed by the same quantity of heat.

Thermal capacity is the most important characteristic in the physics of soil heat so far as we are concerned. It is defined as the heat, in calories, needed to raise the temperature of 1 cm^3 of soil 1°C.

Finally, temperature conductivity is among the thermophysical characteristics of soil, and it characterizes the warming effect of the heat entering the soil. It expresses the combined effect of thermal conductivity (λ) and thermal capacity ($c\rho$)

$$K = \lambda/c\rho$$

We have no instruments for direct measurement of the vertical rise of heat in the upper layers of the soil, so we calculate this important component of the heat balance of the active surface. Analyzing and comparing the different methods of making the calculation, it can be established that today Q_t for soil [2] can be most accurately determined by using Tseytin's method, which uses the following formula to calculate Q_t for the upper 20 cm layer /367

$$Q_t = \frac{c\varrho}{\tau}\left(S_1 - \frac{K}{10} S_2\right) \tag{1}$$

where $c\varrho$ is the thermal capacity; τ is the time interval for which we determine the heat exchange taking place in the soil; S_1 is a function of the change in the heat content of the examined layer; S_2 is a function of temperature changes in individual depths; and K is the temperature conductivity.

Tseytin considered the expression given by Lajhtman:

$$Q_t = \frac{c_1 \varrho_1}{H}\int_0^H [T(z,t) - T(z,0)](H-z)\,dz - \frac{c_1 \varrho_1 K}{H}\int_0^t [T(H,\tau) - T(0,\tau)]\,d\tau \tag{2}$$

where H is the thickness of the examined layer; z is the depth (z = 0 is the soil surface); T is the temperature; t is the time (τ is the change in time). The first integral characterizes the mean heat content for a soil layer of H thickness, and can be determined rather accurately. The second integral depends on the temperature difference between the surface and depth H, calculated for unit distance, that is, it depends on T $\frac{(H,\tau) - T(0,\tau)}{H}$ which also depends on $c_1\ \varrho_1$ K for the layer. Because of the change in λ and K within wide ranges, this integral cannot be determined as accurately as the first integral. Moreover, $\frac{T(H,\tau) - T(0,\tau)}{H}$ is significant for most of the day, and this increases even more the weight of the less reliable part of Eq. (2).

In deriving the Q_t formula, Tseytin started with

$$c_1 \varrho_1 \frac{\partial T}{\partial t} = c_1 \varrho_1 K \frac{\partial^2 T}{\partial z^2} \tag{3}$$

Taking into account physical and mathematical considerations, he arrived at the following expression

$$Q_t = c_1 \varrho_1 \int_0^H [T(z,t) - T(z,0)]\, m(z)\,dz - \frac{c_1 \varrho_1 K}{H-h}\int_0^t [T(H,\tau) - T(h,\tau)]\,d\tau \tag{4}$$

where h is an intermediate level in a layer of H thickness, and m(z) changes its value with depth z as follows

$$m(z) = 1, \text{ ha } 0 < z < h,$$
$$m(z) = \frac{H-z}{H-h}, \text{ ha } h < z < H.$$

This is the formula proposed by Tseytin for calculating the soil heat exchange.

Here the second integral is smaller than the corresponding integral in Eq. (2) because $\frac{T(H,\tau) - T(0,\tau)}{H}$ is larger than $\frac{T(H,\tau) - T(h,\tau)}{H - h}$ which is evident if we consider that the former value is the temperature difference between 0 and H level, while the latter is the temperature difference between an intermediate h and the H level, calculated in unit distances. /368

The calculations made using Eq. (1), or Eq. (4), are very troublesome. Moreover, the uncertainty in determining the second member justified the fact that some authors tried to simplify the formula. Rusin held it possible to ignore the second term in the parenthesis as compared to the first term; that is, he used Eq. (1) in the following form

$$Q_t = \frac{c\varrho}{\tau} S_1 \tag{5}$$

which means he calculated the heat entering the soil with consideration given only to the change in heat content [3].

Numerous measurements and calculations were made [4] to determine the accuracy of these two formulas, and it was found that if we assume the values calculated using Eq. (1) to be conditionally accurate, then the mean relative deviation, which we obtain by using the simplified form of Eq. (1), that is, Eq. (5), approximates in order of magnitude the accuracy with which we now determine the other components of the active layer heat balance. Exceptions are the hours when the heat flux in the soil reverses its signs, but the heat flux is very insignificant in absolute terms [4].

In June 1959, during the Balation research program of the National Meteorological Institute, we measured the heat balance in the area of the Tihany Peninsula. We measured the values of the meteorological elements needed to determine the individual components of the heat balance over a 25-day period, as well as soil temperatures at 0, 5, 10, 20, and 30 cm. We placed our soil thermometers in the 30-40 cm mantle of brown forest loam covering trap-tuff layers, on a southern slope under cultivation (lavender plantation). We continued hourly measurements from 3 June to 27 June excepting when it rained. We determined thermal capacity and soil moisture for soil samples from the 0-5, 5-10, 10-20, and 20-30 cm layers, we used Tseytin's formula, as simplified by Rusin, to calculate heat exchange in the soil. We divided the 30-cm soil segment into 5 and 10 cm thick layers because soil thermal capacity depends on water content which itself changes with the distance of the layer from the surface [5].

We were able to measure individual components of the heat balance without difficulty for 22 of the 25 days of the field measurements without being troubled by lasting rain.

We have included the mean daily behavior of soil heat exchange, as calculated for the 22 days. The mean daily behavior of Q_t (gcal/cm^{2}hr) was:

Hour	Q_t	Hour	Q_t	Hour	Q_t	Hour	Q_t
0–1	–3,0	6– 7	4,6	12–13	6,7	18–19	–6,3
1–2	–3,7	7– 8	8,2	13–14	3,6	19–20	–6,2
2–3	–3,0	8– 9	10,3	14–15	0,8	20–21	–5,5
3–4	–3,3	9–10	11,2	15–16	–3,6	21–22	–4,7
4–5	–1,7	10–11	10,0	16–17	–7,0	22–23	–3,5
5–6	0,4	11–12	9,3	17–18	–7,1	23–24	–3,5

After summing the values obtained for individual time intervals, our results /369 show that during an average day, 65.1 gcal of heat flow from a lccm^2 surface into the deeper layers, and 62.1 gcal of heat flow upwards from the deeper layers through a 1 cm^2 surface. The difference is 3.0 gcal, which means on an average day in June 1959, the soil heat exchange ended with a gain of 3.0 gcal of heat per cm^2 of surface area for warming the deeper layers. The peak in the daily heat exchange occurs between 0800 and 1000. This shift of the peak for heat entering the soil from 1200 to 0800-1000 can be explained by the fact that the steepest temperature gradient develops in the 0-30 cm layer at this time, promoting intensive penetration of heat into the soil.

An interesting phenomenon occurs in the afternoon hours. Although the sun is still shining, the heat flux changes signs. The hourly heat loss between 1600 and 1800 is heaviest in the top layer of the soil.

The development of cloud cover greatly influences the amount of energy from the sun that reaches the surface. Figure 1, in addition to the mean daily behavior of soil heat exchange, also shows the mean daily behavior of the cloud cover. We give the cloud cover on the y-axis in percentages.

Now let us see how the soil manages heat when the sky is clear, cloudy, and overcast. We tabulated the number of days of each for the period, and found that we had had seven days that satisfy clear sky conditions (the 24 hour cloud cover mean was < 20%); 10 days that were cloudy (cloud cover between 21% and 81%), and 5 days that were overcast (cloud cover > 80%). Figure 2 shows the daily behavior of the heat flux and of the cloud cover on these clear, cloudy, and overcast days. We can state that the heat dissipated in the soil is strictly inversely proportional to the cloud cover. On clear, cloudy, and overcast days, the total of Q_t for heat exchanged in the soil was:

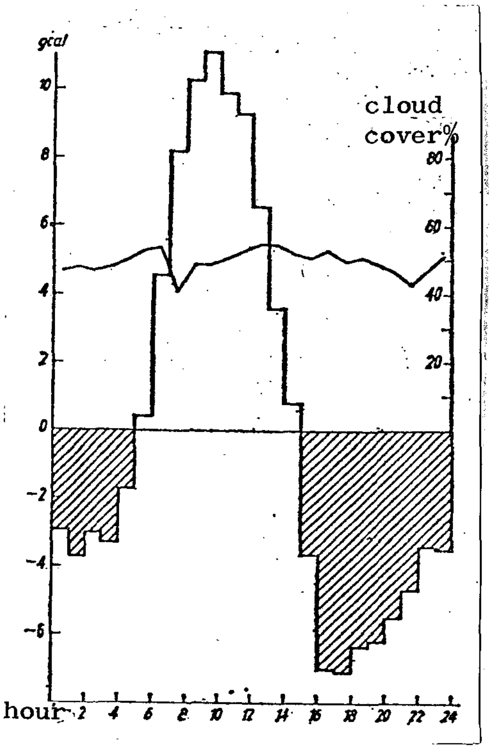

Figure 1. Mean daily behavior of soil heat exchange and of cloud cover, 3-27 June 1959.

	clear	cloudy	overcast
1. heat conducted into the soil	87.6	63.5	44.0 gcal
2. heat dissipated from the soil	79.8	59.4	43.1 gcal

These findings lead to the conclusion that of the energy reaching the surface from the sun on an average, clear, cloudy, and overcast day, 87.6 gcal, 63.5 gcal, and 44.0 gcal respectively, of heat is supplied to the soil, and that 79.8 gcal, 59.4 gcal, and 43.1 gcal, respectively, of heat rise from the soil per 1 cm^2, to raise the surface temperature. This plays a positive role in the active surface heat balance. The daily behavior of the global radiation reaching /370 the surface during the daily behavior of soil heat exchange is directly proportional to the cloud cover, as is shown in Figure 3. A daily mean of 559.4 gcal

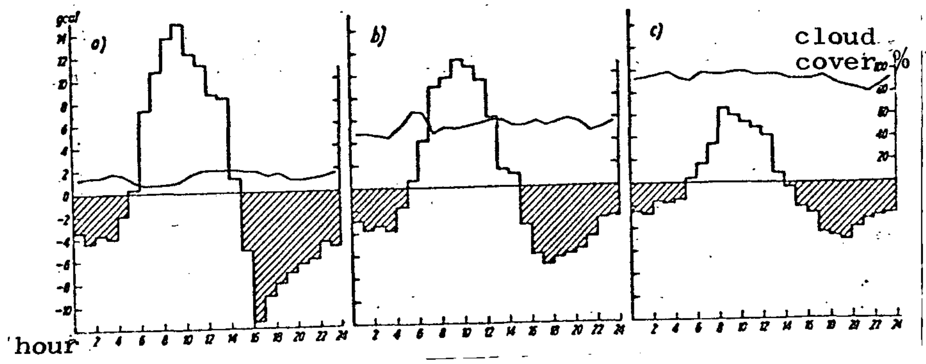

Figure 2. Mean behavior of heat exchange in soil and of cloud cover on (a) clear, (b) cloudy, and (c) overcast days.

of heat were received per cm^2 of surface from global radiation during the measurement period, with the figures for an average clear, cloudy, and overcast day 619.9, 568.1, and 418.1 gcal.

The following expression

$$R + L \cdot E + Q_l + Q_t = 0$$

used to determine the surface soil heat balance is called the basic heat balance equation. In addition to Q_t, we already have discussed [6, 7] the determination of L • E and Q_l. Zs. Tarkanyi determined R as part of our heat balance measurements. Using his data, we can compare the values of R, the radiation coefficient, Q_l, the air conductivity coefficient, Q_t, the soil heat flux coefficient, and L • E, the coefficient of aggregate change. This paper cannot of course, explain in detail or discuss any changes in the values of the coefficients that make up the basic heat balance equation. We would like to present a more detailed study of the heat balance at another time. Here we show only in rough outline the percentage distribution of the individual coefficients for an average day in June.

According to our calculations, in June 1959, the radiation balance (R) was positive from 0500 to 1900, and during that period 1 cm^2 of surface received 397.2 gcal (Q_t) of heat. Of this, 43.7 gcal (Q_t) penetrated into the deeper layers of the soil, and evaporation and turbulent heat exchange transported 353.1 gcal ($L \cdot E + Q_t$) into the upper layers of the atmosphere.

Thus, of the energy originating from solar radiation and providing a surface gain, 11% went to warm the soil, 89% to evaporation and warming the atmosphere. On the other hand, the radiation coefficient was negative between 1900 and 0500; that is, the surface was losing heat by radiation. To replace this heat loss, the surface received 37.9 gcal from the soil by conduction, with the air conduction coefficient, positive during the night hours, contributing. In general, evaporation continued during the night, adding to the heat loss in the surface energy balance. We found dew in only a few cases, but when we did the surface received additional heat because dew formation is accompanied by liberation of heat.

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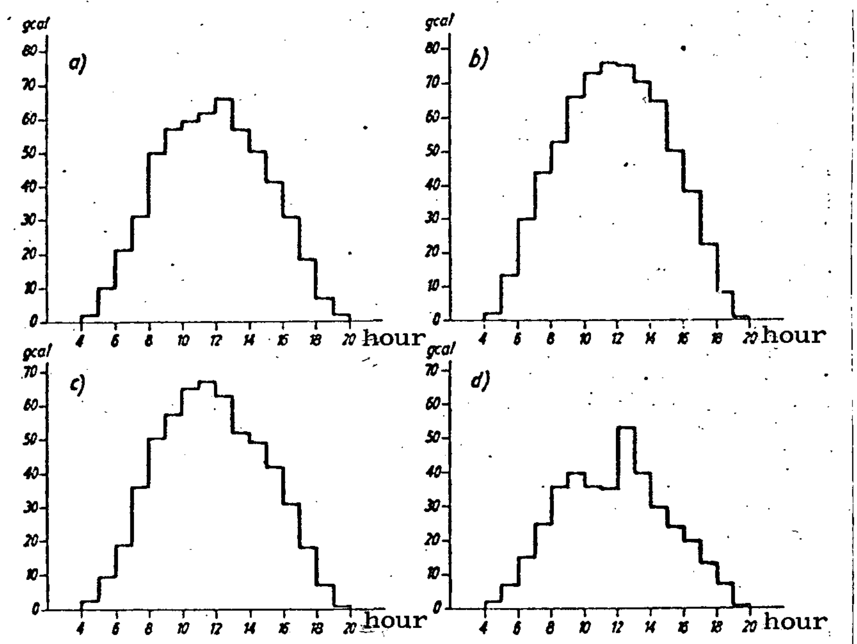

Figure 3. Mean daily behavior of global radiation (a) in the period June 3-27, 1959; (b) on clear days, (c) on cloudy days, and (d) on overcast days.

From what has been said thus far, the fact seems to be evident that at those stations where measurements of the heat balance were made determination of the soil heat flux is required. But there are other goals to be reached, including as accurate a determination of the value of this coefficient as possible. Of interest to use, and primarily from an agricultural standpoint, is how different soils deal with the heat they receive. Of course, investigations of this type require that we measure characteristic soil types. Finally, our investigations may serve as a guide in the work of compiling a soil temperature map of Hungary. A map such as this now is very timely and should be compiled in the near future.

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Translated for the National Aeronautics and Space Administration under contract no. NASw 2038 by Translation Consultants, Ltd., 944 South Wakefield Street, Arlington, Virginia 22204.